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PHYSICAL PROPERTIES AND ENERGY DISTRIBUTION OF GULF STREAM EDDI--ETC(U)

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E. Khedouri
W. Gemmill

U.S. NAVAL OCEANOGRAPHIC OFFICE
WASHINGTON, D.C. 20373

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Temperature, salinity, sound speed, gradient currents and energy distribution of two Gulf Stream eddies are described. The eddies are different in rotation and size, but are similar in energy distribution and ratio of kinetic to potential energy. This ratio is equal to the eddy number and Rossby number.

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ABSTRACT

Selected physical properties including temperature, salinity, sound speed, gradient currents and energy distribution of two Gulf Stream eddies are described. These eddies of different rotation and size, exhibited certain similarities in energy distribution and in the ratio of kinetic to available potential energy. It is shown theoretically, that under geostrophic assumption, this ratio is approximately equal to the eddy number and the Rossby number.

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I. INTRODUCTION

The objective of this paper is to describe and compare selected properties of two Gulf Stream eddies, one cyclonic and one anticyclonic, which were extensively surveyed by the U.S. Naval Oceanographic Office from ships and aircraft. Ship surveys included S/T/D (salinity, temperature, depth) measurements at approximately 20 kilometers spacing across the eddies, providing good data for calculating energy distribution within the eddies. Location of the two eddies and the S/T/D stations are shown in figure 1.

Energy in the ocean consists of kinetic energy (KE) and potential energy (PE). If the ocean density was horizontally stratified and statically stable, PE would not be available for conversion to KE and the minimum PE state would exist. However, the ocean is not horizontally stratified everywhere (although it is generally statically stable) and, therefore, some of the PE is available to be released into motion (KE) through redistribution of mass to the minimum PE state. PE which is available for conversion to KE is called the available potential energy (APE). For an eddy, the minimum PE state is that of the surrounding stationary water; Sargasso Sea for the cyclonic and Slope Water for the anticyclonic eddy. The APE is the difference between the PE within the eddy and the PE of the surrounding water. The concepts of energy used here are similar to those developed

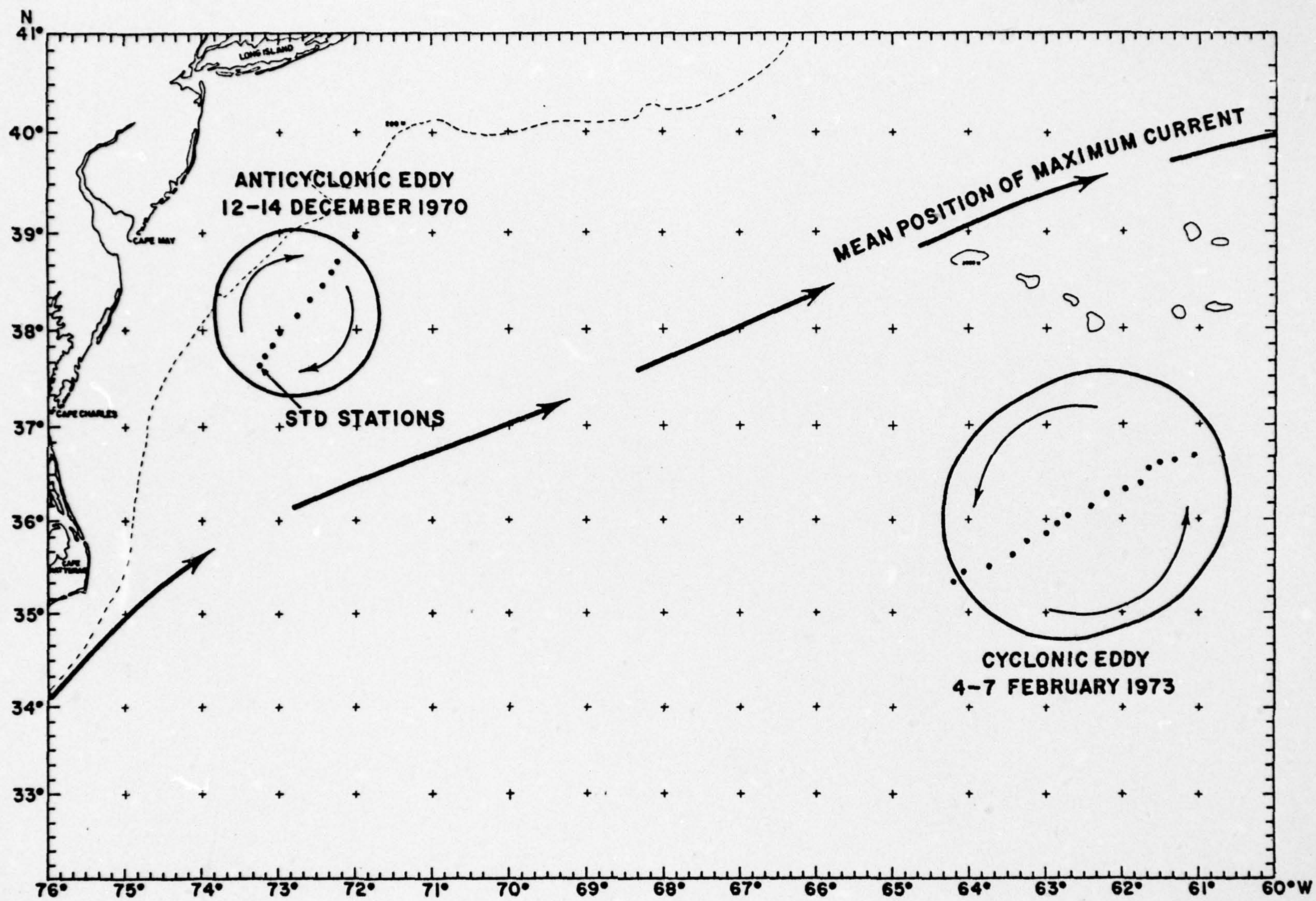


Figure 1 - Location of the eddies and the S/T/D stations

by Barrett (1971) and Wright (1972). Understanding energy distribution within eddies and the conversion rate of APE to KE may be useful in predicting eddy decay rates and life cycles.

II. THERMOHALINE STRUCTURE OF THE EDDIES

Vertical temperature and salinity sections of the two eddies are shown in figures 2 and 3. Note the radius of the cyclonic (cold) eddy is nearly twice that of the anti-cyclonic (warm) eddy. The differences between cyclonic and anticyclonic eddies are evident by inspection of their thermohaline structure. The anticyclonic eddy is a relatively shallow feature extending from surface to about 1000 meters. It is composed of warmer and more saline water than the surrounding Slope Water. The cyclonic eddy, however, extends to below the maximum extent of the data which was 2500 meters. It is composed of colder and less saline water than the surrounding Sargasso Sea. Because cyclonic eddies are denser than the surrounding water, they gradually sink at a rate of 1.6 meters a day (Parker 1971), and in a typical case within two or three months after formation the eddy is completely submerged and cannot be detected from surface temperature or salinity. Warm eddies would be expected to rise, because they are less dense than the surrounding water, but because they are susceptible to air-sea interactions, the rate of rise is difficult to measure.

ANTICYCLONIC
EDDY

CYCLONIC
EDDY

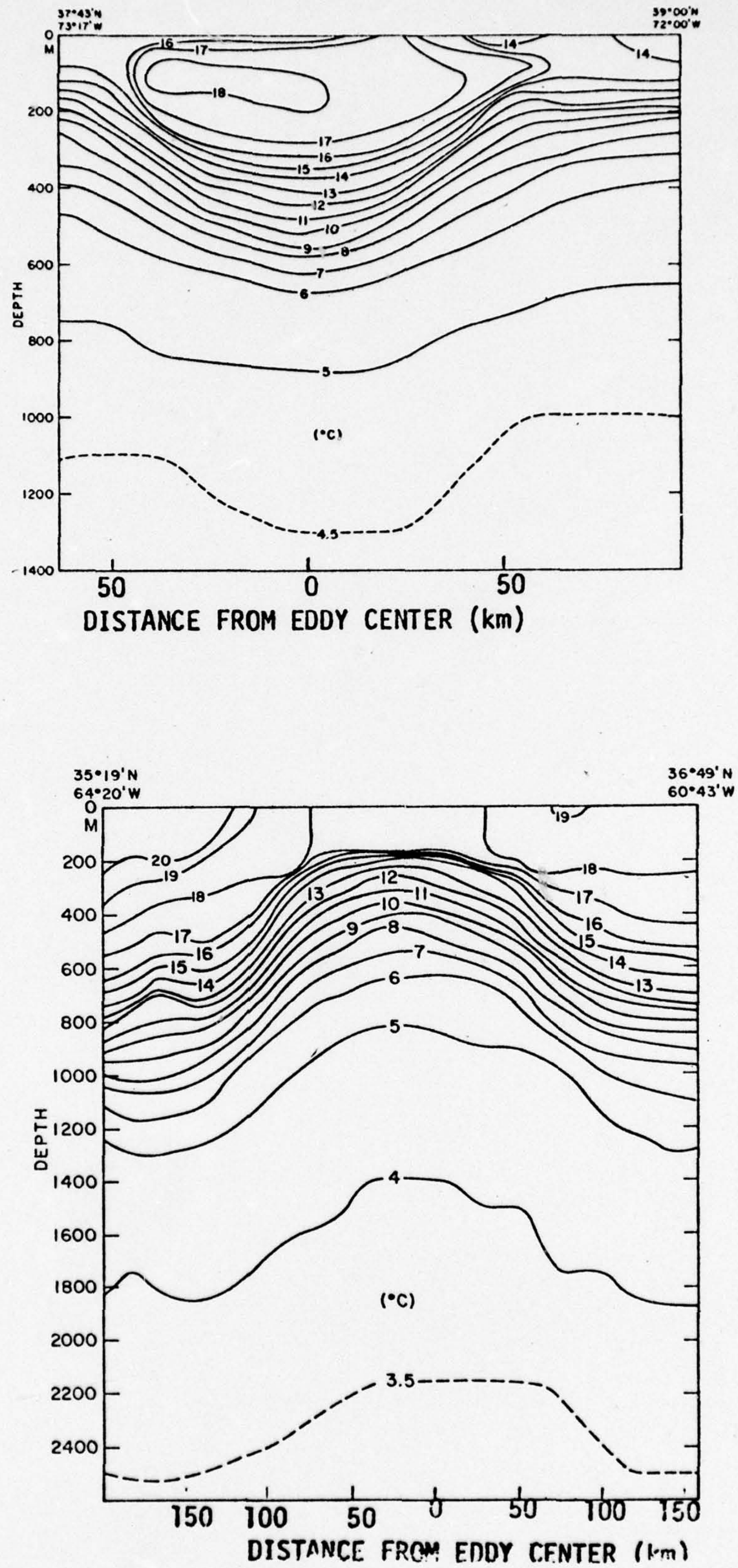


Figure 2 - Vertical temperature sections

ANTICYCLONIC
EDDY

CYCLONIC
EDDY

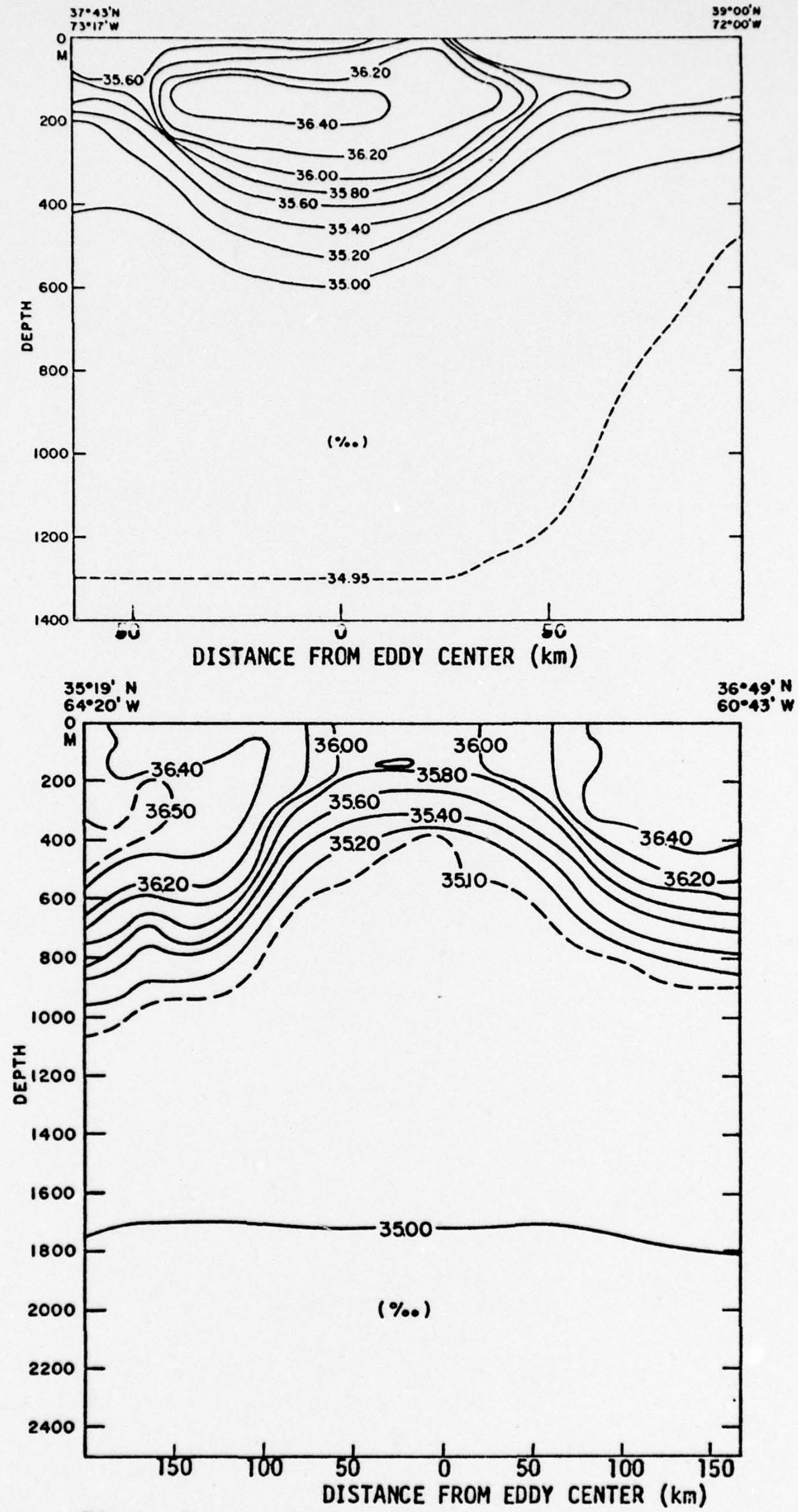


Figure 3 - Vertical salinity sections

III. SOUND SPEED STRUCTURE

Vertical sound speed sections of the two eddies shown in figure 4 are nearly the inverse of one another. Sound speeds were computed from depth, temperature, and salinity data using Wilson's (1960) equations. The isovels reflect changes in temperature and salinity across the eddies. Relative to the surrounding waters, sound speeds are lower in cold eddies and higher in warm eddies.

Two parameters useful in the description of vertical sound speed structure are the sonic layer depth (SLD) and the deep sound channel (DSC) axial depth. SLD is defined as the depth of maximum sound speed above the axis of the deep sound channel, and the deep sound channel axial depth is defined as the depth of the minimum sound speed. SLD is strongly dependent upon air-sea interactions exhibiting both spatial and temporal changes. DSC axis on the other hand is dependent only on large scale oceanographic circulation patterns. It may change abruptly across an oceanic front or an eddy.

DSC axial depth from the Sargasso Sea to the center of the cold eddy decreases from 1300 meters to 700 meters and the corresponding sound speed along the DSC axis decreases from 1493 m sec^{-1} to 1484 m sec^{-1}. A similar sound speed structure has been computed in a cold eddy by Vastano and Owens (1973). In contrast, the DSC axis for the warm eddy deepens from 550 meters in Slope Water to 750 meters at the center of the eddy, and its accompanying sound speed increases from 1481

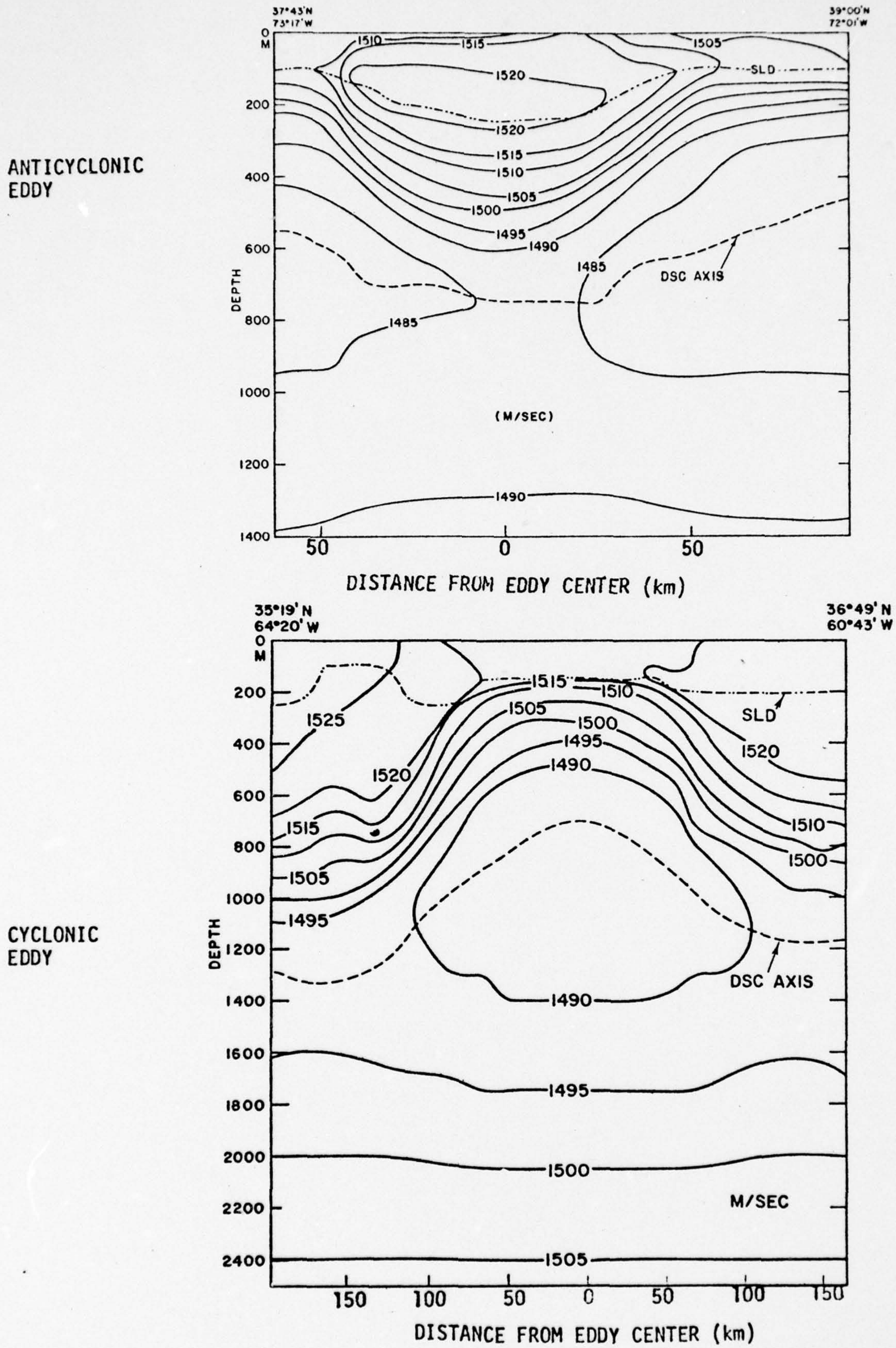


Figure 4 - Vertical sound speed sections

to 1485 m sec^{-1}. In both eddies the sound speed changes along the DSC axis are relatively small (5-8 m sec^{-1}), however, sound speed changes across the eddies at a constant depth within the upper 600 meters may be as large as 30 m sec^{-1}. Vertical extent or thickness of DSC is increased by the cold eddy and decreased by the warm eddy.

SLD in the warm eddy is at 250 meters corresponding to the nearly isothermal structure to that depth. SLD in the surrounding Slope Water is at 100 meters. The reason for the deeper SLD within the eddy is not clearly understood. It appears that fall cooling and horizontal mixing resulted in more intense cooling within the eddy than in the surrounding Slope Water. Saunders (1971) has reported a similar observation for a warm eddy studied during the same season in 1969. SLD of the cold eddy is almost constant at 150 meters and the SLD in the surrounding Sargasso Water varies between 50 and 200 meters. The reason for small SLD difference between the eddy and the surrounding water is that this eddy has sunk to a point where its influence on SLD is insignificant.

IV. DYNAMIC TOPOGRAPHY AND GRADIENT CURRENTS

Density difference between an eddy and its surrounding water results in distortion of dynamic topography. The anti-cyclonic eddy, which is composed of less dense water than the surrounding Slope Water, forms a bulge in the ocean surface with a maximum height difference of 26 dynamic cm. The cyclonic eddy which is composed of denser water than the surrounding

Sargasso Sea, forms a depression with a maximum height difference of 92 dynamic cm. By way of comparison the height difference across the Gulf Stream is on the order of 100 dynamic cm.

Gradient currents and volume transport associated with these dynamic anomalies are shown in figure 5. Both eddies show a ring like circulation with maximum current near the surface and midway between the perimeter and the center of the eddy. Currents were calculated by correcting geostrophic flow for curvature using the gradient wind equation. The equation is derived by balancing Coriolis, pressure gradient, and centrifugal forces. For a cyclonic eddy, the corrected current velocity (Vgr) is:

$$Vgr = -\frac{Rf}{2} + [\frac{R^2f^2}{4} + f|Cg|R]^{1/2} \qquad (cm\ sec^{-1})$$

and for an anticyclonic eddy the corrected current velocity is

$$Vgr = +\frac{Rf}{2} - [\frac{R^2f^2}{4} - f|Cg|R]^{1/2} \qquad (cm\ sec^{-1})$$

where R = radius of curvature (cm)

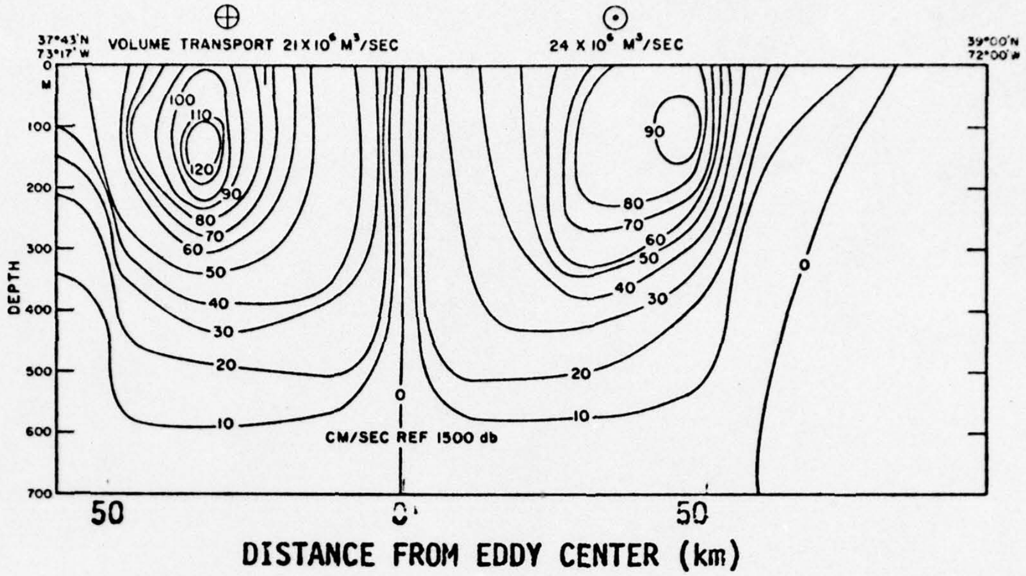
f = Coriolis parameter (sec^{-1})

Cg = geostrophic current speed (cm sec^{-1})

In a cyclonic eddy, the gradient current is less than geostrophic and in anticyclonic eddy it is greater than geostrophic.

Note that in an anticyclonic eddy, for any given radius, there is a limit to the magnitude of pressure gradient force

ANTICYCLONIC EDDY



CYCLONIC EDDY

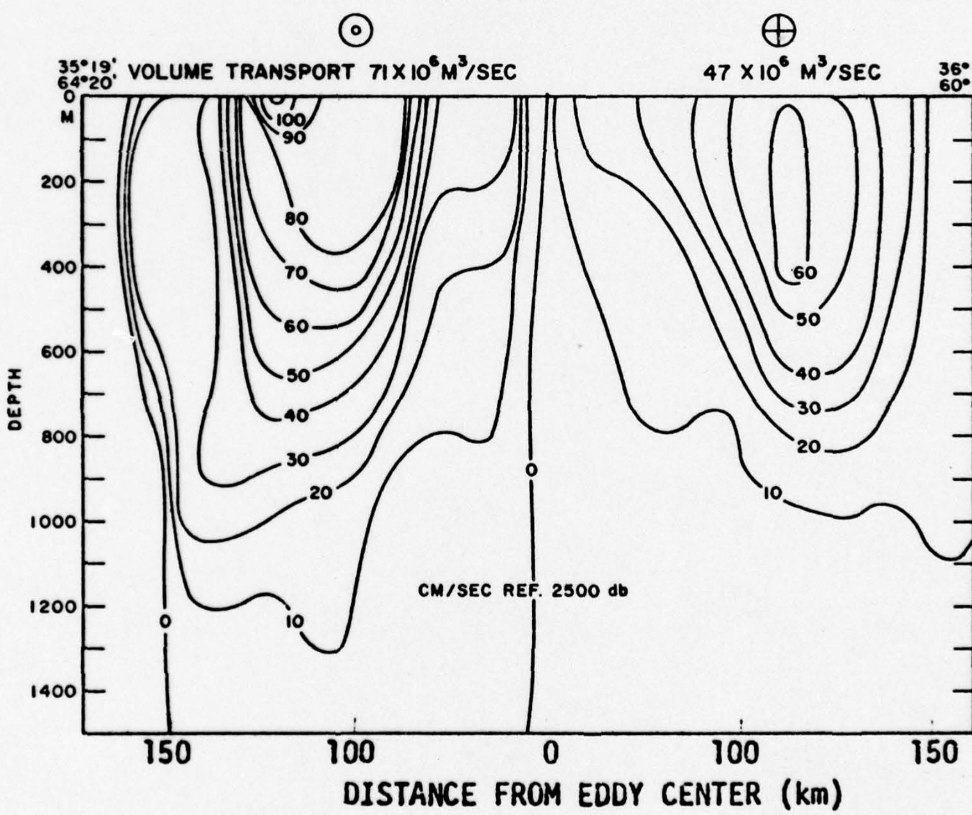


Figure 5 - Gradient currents and volume transport

beyond which the forces will not be in balance. That is, the term under the square root of the gradient wind equation must be positive, otherwise the eddy becomes unstable and will eventually collapse.

V. ENERGY DISTRIBUTION

As a cold eddy sinks or a warm eddy rises to its equilibrium isopycnal level, the available potential energy is transformed to kinetic energy, which in-turn is dissipated by friction. APE can also be destroyed or generated by differential heating or cooling between the eddy and the adjacent ocean.

APE per unit area is the difference between PE at a point inside the eddy and the PE of stationary surrounding waters. It is defined by:

$$APE_A = \int_Z^0 g\ Z\ (\rho-\rho')\ dZ \quad (\text{ergs cm}^{-2})$$

where Z = depth (cm)

g = gravity (cm sec^{-2})

ρ = density inside the eddy (gm cm^{-3})

ρ' = density outside the eddy (gm cm^{-3})

The total APE is obtained by horizontal summation of APE_A over the area of the eddy.

Assumptions involved in these procedures (Barrett, 1971) are: (1) the eddy is circular, (2) potential energy change in the surrounding water resulting from the collapse of the eddy is negligible, (3) hydrographic sections must pass

through the center of the eddy and (4) lateral translations of the entire eddy which may affect station positions relative to the center of the eddy are ignored.

Aerial surveys with infrared thermometer (ART) and air-borne expendable bathythermographs (AXBT) revealed that these eddies are slightly elliptical. To minimize the error caused by the eddies not being circular (1st assumption), horizontal energy summation was performed by considering the eddy as being composed of two semi-circles of different diameters and integrating energy per unit area over the area in each half of the eddy. The second assumption is realistic for a single eddy since the volume of an eddy is small relative to surrounding waters. The third and fourth assumptions were justified for the two eddies because their centers were located by aerial surveys just prior to ship surveys with S/T/D stations. The S/T/D survey was completed within 2 days for the cyclonic and 4 days for the anticyclonic eddy, and the lateral translation for the duration of the survey was relatively small (10 km for anticyclonic eddy and 30 km for cyclonic eddy) when compared to eddy diameters.

KE per unit area is defined by:

$$KE_A = 1/2 \int_0^Z \rho v^2 \, dZ \ (\text{ergs cm}^{-2})$$

where v = current velocity (cm sec^{-1})

Total KE was calculated by horizontal summation of KE_A over the area of the eddy. As with APE, the summation was

performed over the two halves of the eddies sepearately, to minimize the error caused by the eddies not being circular.

Horizontal distribution of APE and KE per unit area (figure 6) shows that the maximum KE occurs midway between the center and the perimeter of the eddy, corresponding to the location of maximum current velocity, and the maximum APE occurs at the center of the eddies corresponding to the maximum dynamic height anomaly.

Vertical distribution of APE and KE (figure 7) shows that maximum KE occurs at the surface in the cyclonic eddy and at 150 meters in the anticyclonic eddy. Maximum APE in the anticyclonic eddy occurs at 300 meters and in the cyclonic eddy at 1000 meters. A similar distribution of energy was reported by Barrett (1971). The oscillations observed in the APE distribution of the cold eddy are not clearly understood.

Total APE of the cyclonic eddy (referenced to 2500 meters) was 3.0×10^{24} ergs and total KE was 8.5×10^{22} ergs. For the anticyclonic eddy the energies (referenced to 1500 meters) were 1.1×10^{23} ergs for APE and 5.9×10^{21} for KE. Ratios of KE to APE were 1:18 for anticyclonic and 1:35 for cyclonic eddy. Saunders (1971) calculated a ratio of 1:30 for an anticyclonic and Wright (1970) estimated the ratio for the entire ocean to range between 1:10 and 1:50.

VI. EDDY NUMBER AND RATIO OF KE TO APE

A dimensionless number called the eddy number (N_E) defined by Wilkinson (1972) was found to be surprisingly constant for twelve randomly chosen eddies of various sizes.

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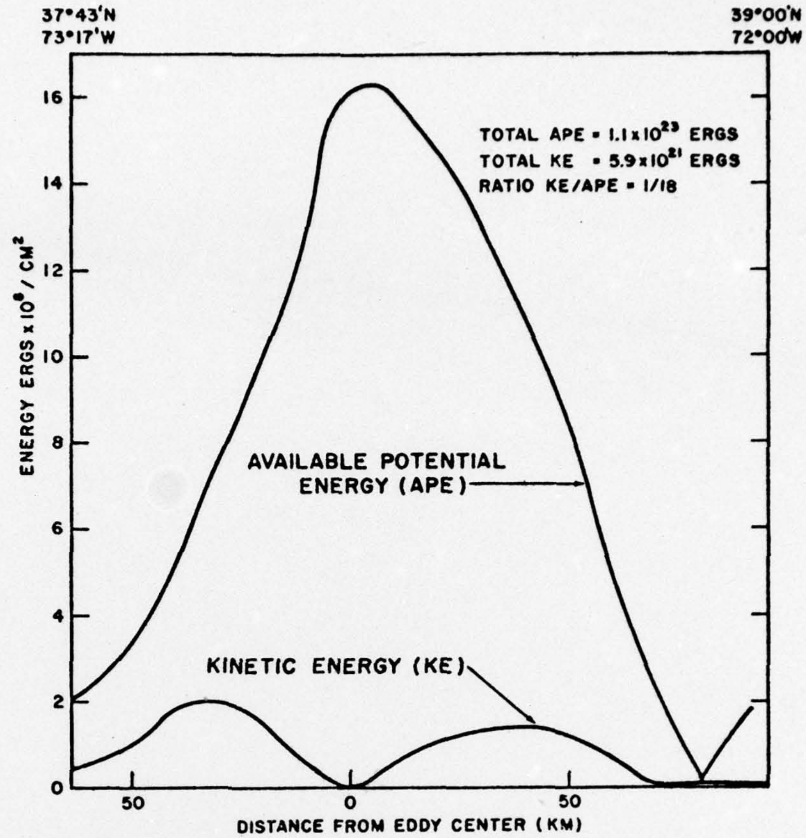
EDDY



CYCLONIC
EDDY

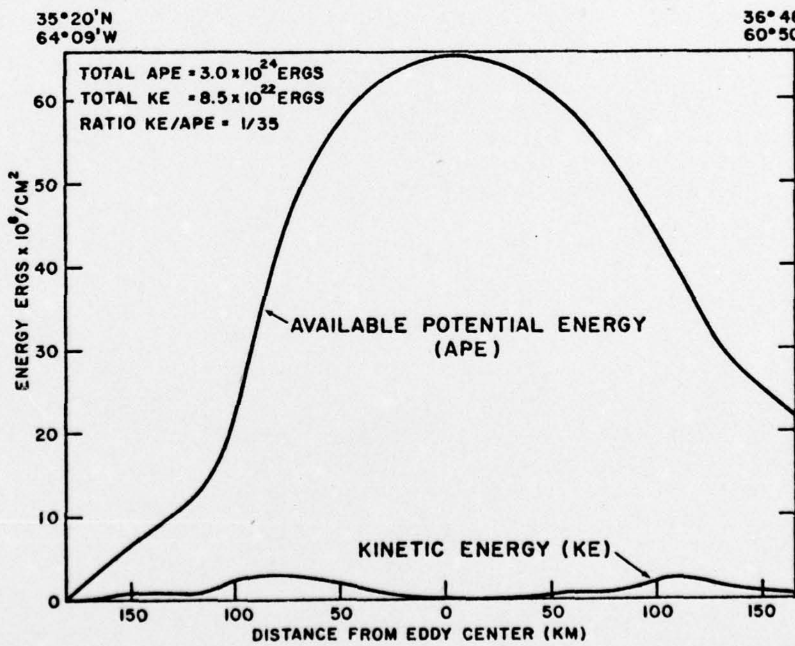
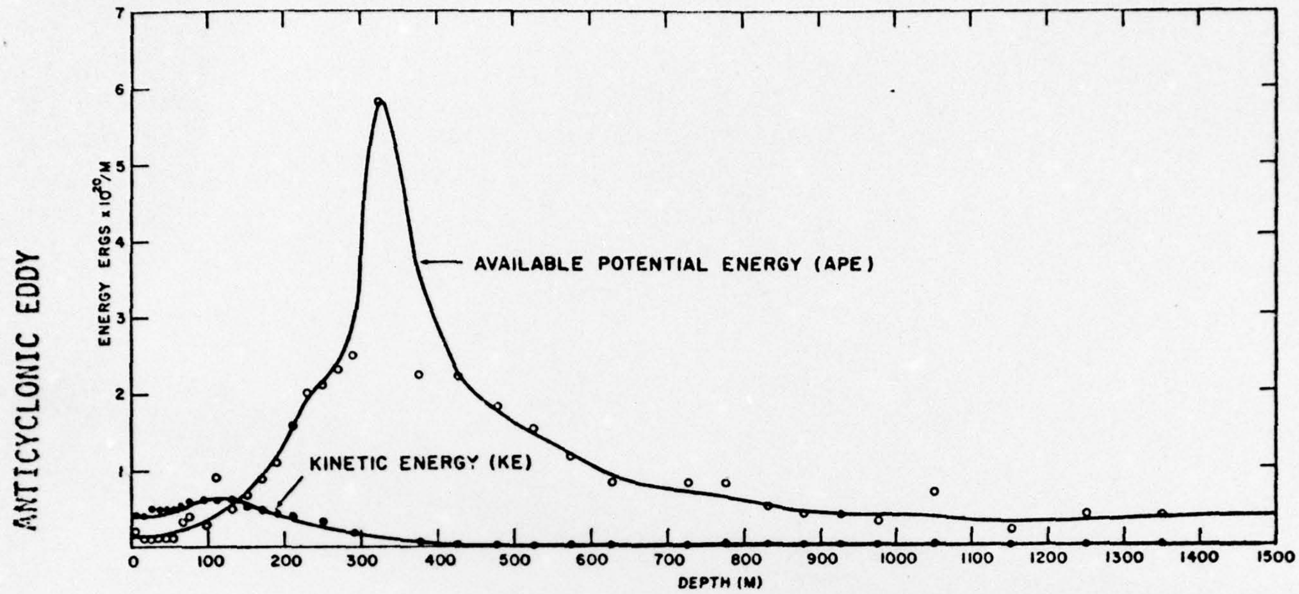


Figure 6 - Horizontal distribution of APE and KE



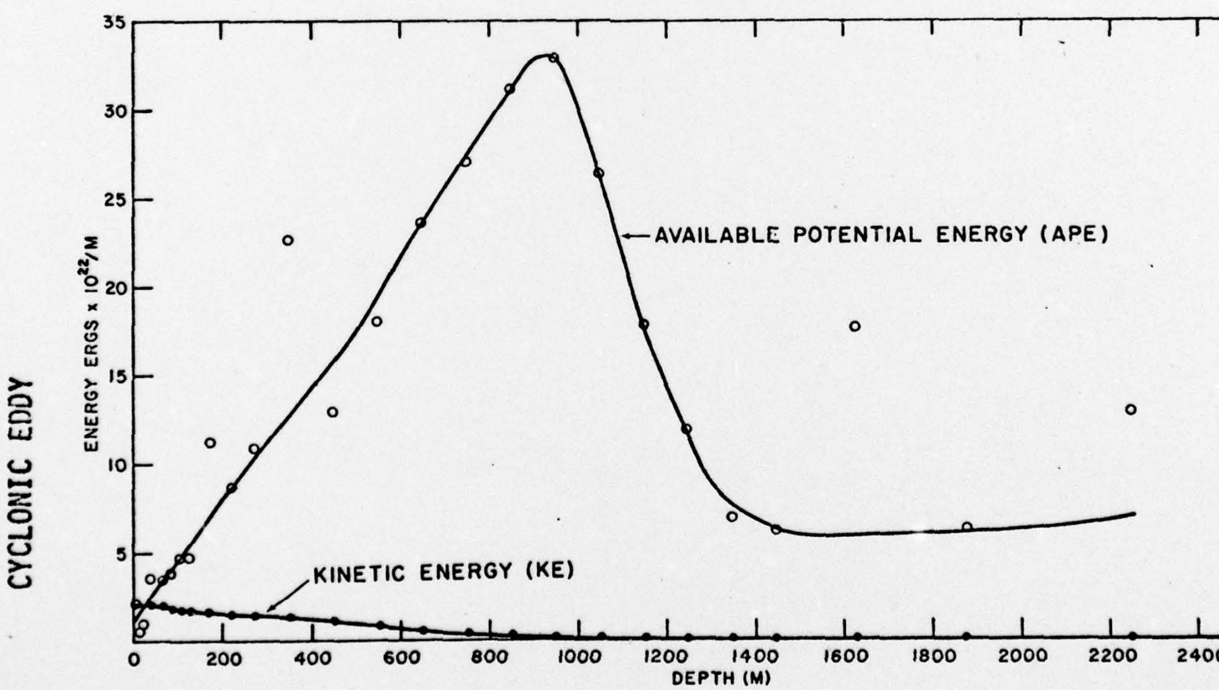


Figure 7 - Vertical distribution of APE and KE

The eddy number is derived assuming geostrophic motion, hydrostatic equilibrium and solid eddy rotation. Assuming geostrophic circulation N_E is identical to Rossby number.

The eddy number is given by:

$$N_E = \frac{g\Delta h}{(Rf)^2}$$

where Δh = geometric height of the eddy relative to surrounding ocean (cm)

g = acceleration of gravity (cm sec^{-2})

R = eddy radius (cm)

f = Coriolis parameter (sec^{-1})

Mean value for the eddy number calculated by Wilkinson for twelve eddies, was 0.036. The eddy numbers for the two eddies described here were 0.038 for the cyclonic and 0.033 for the anticyclonic eddy. The ratios of KE/PE for the two eddies were 0.028 and 0.055 respectively. If geostrophic currents rather than gradient currents were used for KE calculation, the KE of the warm eddy would be decreased and KE of the cold eddy would be increased. When this was done, the ratio of KE/APE for both eddies was calculated to be 0.04. This suggested that there may be some relation between the eddy number and KE/APE ratio.

Derivation of N_E is presented in Appendix A. Making the same assumptions, geostrophic flow, hydrostatic equilibrium and solid rotation, the ratio of KE/APE is derived in Appendix B and it is shown theoretically that $\frac{KE}{APE} \approx N_E$.

VII. SUMMARY AND RECOMMENDATIONS

Two Gulf Stream eddies described here, produce thermohaline, current, and sound speed characteristics quite different from those found in their surrounding waters. Some of these characteristics and differences are summarized in the following three tables.

Table 1 summarizes temperature, salinity and sound speed at 200 meter and 400 meter depths of the eddies and the surrounding waters. Table 2 summarizes the SLD, DSC axis, and temperature, salinity and sound speed along the DSC axis. Finally, Table 3 summarizes some of the dynamic characteristics of the two eddies.

TABLE 1

Comparison of selected physical properties at 200 m and 400 m depths

Parameter depth	Temperature °C 200m	Temperature °C 400m	Salinity °/oo 200m	Salinity °/oo 400m	Sound Speed m sec^{-1} 200m	Sound Speed m sec^{-1} 400m
Cyclonic Eddy (center)	13.8	9.0	35.66	35.18	1507.8	1493.5
Surrounding Water (Sargasso Sea)	19.9	18.1	36.39	36.51	1526.8	1525.1
Difference	-6.1	-9.1	-0.73	-1.33	-19.0	-31.6
Anticyclonic Eddy (center)	18.0	13.2	36.46	35.62	1521.5	1508.1
Surrounding Water (Slope Water)	10.8	5.9	35.13	34.96	1494.6	1481.2
Difference	+7.2	+7.3	+1.33	+0.66	+26.9	+26.9

TABLE 2

Comparison of some specific eddy parameters

	Depth of DSC Axis (m)	Temp°C @ DSC axis	Sal °/∘∘ @ DSC axis	Sound Speed m sec^{-1} @ DSC axis	SLD m
Cyclonic Eddy	700	5.4	35.02	1484.3	150
Surrounding Water	1300	5.0	35.02	1492.7	200
Difference	-600	+.4	0	-8.4	-50
Anticyclonic Eddy	750	5.4	34.95	1485.0	240
Surrounding Water	550	5.2	34.97	1480.9	100
Difference	+200	.2	-.02	+4.1	140

TABLE 3

Summary of dynamic eddy characteristics

Type	Diameter (km)	Height Anomally cm	APE ergs	KE ergs	KE/ APE	N_E
Cyclonic Eddy	350	92	3.0×10^{24}	8.5×10^{22}	0.028	0.038
Anticyclonic	180	26	1.1×10^{23}	5.9×10^{21}	0.055	0.033

Because Gulf Stream eddies produce significant changes in the environment and because they are very large and persistent, their life cycles and decay rates must be studied. This is particularly important for cyclonic eddies because they slowly sink below the surface and are not detectable from surface temperatures. One way to predict their decay rate and life expectancy is to determine the conversion rate of APE to KE. To accomplish this, a cold eddy should be surveyed with S/T/Ds several times at approximately three month intervals. The rate of APE decrease at these periods should provide a good indication of the eddy life expectancy.

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APPENDIX A

FORMULATION OF THE EDDY NUMBER

In Wilkinson's short note (1972), the concept of the eddy number (N_E) was developed. The N_E is identical to Rossby number (R_O) for geostrophic circulation. Derivation of N_E is presented here for completeness.

An eddy is considered as a rotating cylinder in a stationary ocean. The motion is assumed to be in hydrostatic equilibrium and in solid rotation. For this type of motion, the flow is the balance of the pressure gradient ($\frac{1}{\rho}\frac{\partial p}{\partial r}$), Coriolis ($fv$), and centrifugal ($v^2/r$) forces and is given by:

$$\frac{1}{\rho}\frac{\partial p}{\partial r} = fv + \frac{v^2}{r} \qquad (1a)$$

where ρ = density (gm cm^{-3})

v = current velocity (m sec^{-1})

r = radial distance to eddy center (cm)

f = Coriolis parameter (sec^{-1})

p = pressure (dynes cm^{-2})

hydrostatic equilibrium allows (1a) to be written as:

$$g\frac{\partial h}{\partial r} = \frac{v^2}{r} + fv \qquad (2a)$$

Further, if $D = g\Delta h$, where D is the difference between the dynamic height in the center of the eddy and the dynamic height of the stationary adjacent water, and if the eddy is in solid rotation so that $V = \Omega R$, then (2a) can be written as:

$$\frac{D}{R} \simeq \Omega^2 R + f\Omega R \qquad (3a)$$

where: R = radius of the eddy (cm)

Ω = eddy rotation rate (sec^{-1})

The eddy number is obtained by dividing (3a) by Rf^2:

$$N_E = \frac{D}{f^2R^2} = \frac{\Omega^2}{f^2} + \frac{\Omega}{f} \tag{4a}$$

It is this number (N_E) that is conservative for a wide range of eddies. Also, because the Coriolis parameter is an order of magnitude larger than the the rotational rate of oceanic flows ($f \gg \Omega$), (4a) can be further simplified (geostrophic approximation).

$$N_E = \frac{D}{f^2R^2} \simeq \frac{\Omega}{f} \equiv R_o \tag{5a}$$

Therefore, for geostrophic motion, N_E is identical to R_o.

APPENDIX B

THE KE/APE RATIO

In order to obtain an order of magnitude approximation for the KE to APE ratio of an eddy, similar simplifying assumptions as in Appendix A are made. That is the eddy is axially symmetric, geostrophic, and in hydrostatic equilibrium. The total KE of an eddy is then defined by (in cylindrical coordinates):

$$KE = \frac{\pi}{g} \int_{p}^{o} \int_{o}^{R} v^2 \, dp \, rdr \qquad (1b)$$

where v = current velocity (cm sec^{-1})

g = acceleration of gravity (cm sec^{-2})

r = radial distance to eddy center (cm)

p = depth measured in terms of pressure (dynes cm^{-2})

R = radius of the eddy (cm)

using the geostrophic approximation (scaled for an eddy):

$$V = \frac{D}{fR} \qquad (2b)$$

where D = $g\Delta h$ the dynamic height of difference between the center of the eddy and the surrounding stationary ocean

f = Coriolis parameter (sec^{-1})

then the geostrophic KE can be rewritten as:

$$KE = \frac{\pi}{2g} \, v2R2p = \frac{\pi D^2 p}{2gf^2} \qquad (3b)$$

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